Revisiting sites of the South Pole Queen Maud Land Traverses in East Antarctica: Accumulation data from shallow firn cores

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[1] Ground-based accumulation measurements are scarce on the high East Antarctic plateau, but highly necessary for model validation and the interpretation of satellite data for the determination of Antarctic mass balance. Here, we present accumulation results obtained from four shallow firn cores drilled in the Antarctic summer season 2007/2008. The cores were drilled along the first leg of the Norwegian-US IPY traverse through East Antarctica, visiting sites like Plateau Station and Pole of Relative Inaccessibility that have been covered by the South Pole Queen Maud Land Traverses (SPQMLT) in the 1960s. Accumulation has been determined from volcanic chronology established from the conductivity records measured by dielectric profiling (DEP). The Tambora 1815/unknown 1809 double peak is clearly visible in the conductivity data and serves as a reliable time marker. Accumulation rates averaged over the period 1815–2007 are in the range of 16 to 32 kg m^{-2} a^{-1}, somewhat lower than expected from the SPQMLT data. The spatial pattern is mainly influenced by elevation and continentality. Three of the firn cores show a decrease of more than 20% in accumulation for the time period 1815–2007 in relation to accumulation rates during the period 1641–1815. The spatial representativity of the firn cores is assessed by ground-penetrating radar, showing a rather smoothly layered pattern around the drill sites. Validation of the DEP results is utilized by comparison with chemistry data, proving the validity of the DEP method for dating firn cores. The results help understanding the status of the East Antarctic ice sheet and will be important for e.g. future model-derived estimates of the mass balance of Antarctica.

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1. Introduction

[2] Sea-level rise has been a much debated issue in recent climatological studies with predictions ranging from 0.18 to 0.59 m by the end of the 21st century relative to 1980–1999 [*Intergovernmental Panel on Climate Change* (*IPCC*), 2007]. However, studies showed that recent observational data are on the upper margin or even exceeding former model predictions [*Rahmstorf et al.*, 2007]. One of the main uncertainties arises from the still unknown contribution of the (East) Antarctic ice sheet [*Alley et al.*, 2005]. Hence, assessing the mass balance and surface-mass balance of the (East) Antarctic ice sheet has been a major concern of recent studies [*Vaughan et al.*, 1999; *Giovinetto and Zwally*, 2000; *Arthern et al.*, 2006; *van de Berg et al.*, 2006]. Yet despite the promising results of new satellite techniques [*Velicogna and Wahr*, 2006; *Chen et al.*, 2006] today even the sign of the East Antarctic contribution to sea-level rise remains under debate. For example, *Davis et al.* [2005] report an elevation rise/growth of East Antarctica from satellite data between 1992 and 2003 primarily attributed to short term snowfall variability, while *Monaghan et al.* [2006] using a combination of ice core data and meteorological modeling found no long term snowfall trends on the East Antarctic ice sheet since 1957. A new study by *Helsen et al.* [2008] underlines that even insignificant trends in accumulation have an impact on elevation change and need to be taken into account when assessing ice-mass budget.

[3] Several approaches of constraining the mass balance of Antarctica utilize interpolation of accumulation rates obtained from field data such as firn cores, snow pits or stake readings, sometimes with background fields from satellite data for control of the interpolation scheme

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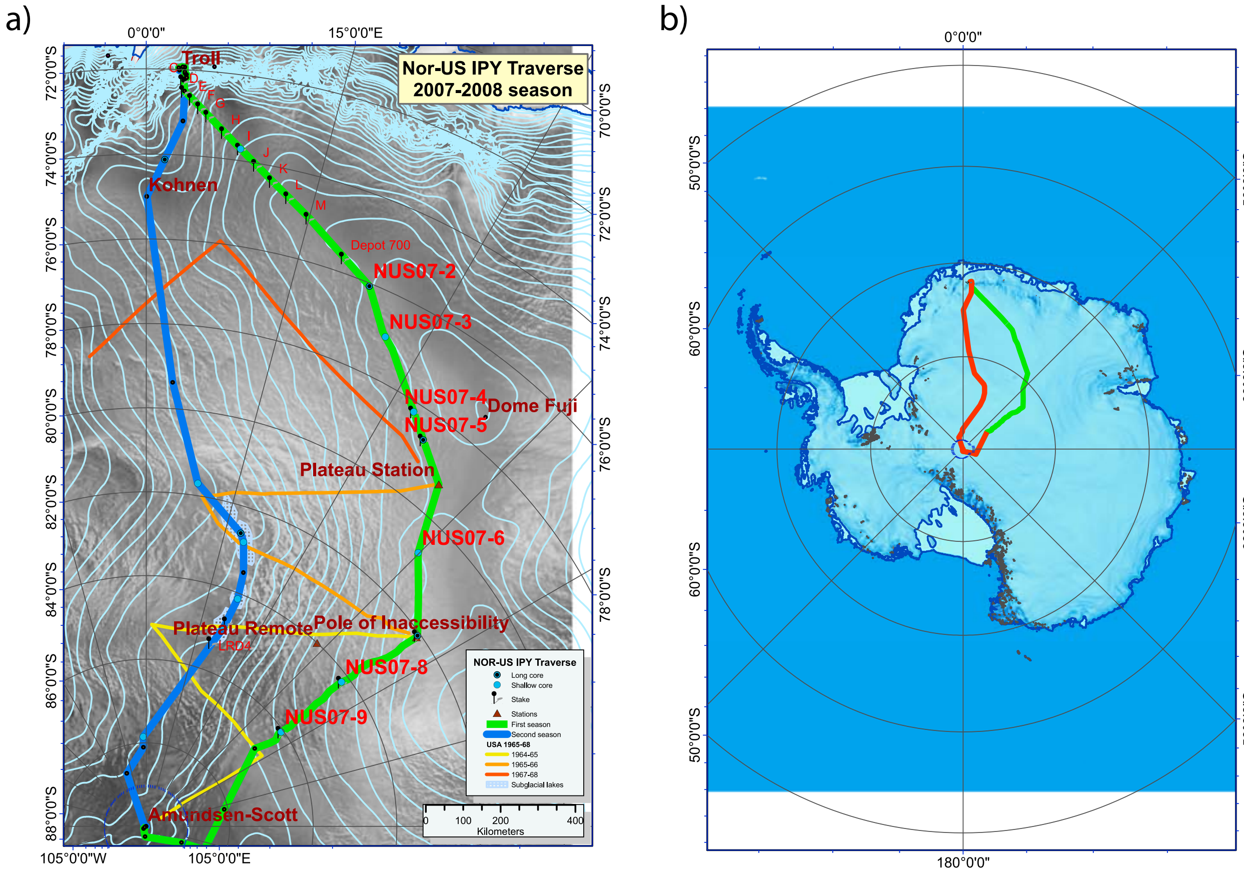

Figure 1. (a) Map of the traverse route 2007/2008 (green line) and 2008/2009 (blue line) with drill sites from the first leg 2007/2008 indicated (NUS07-X). The points C–M mark points of previous Norwegian traverses [*Winther et al.*, 1997, 2002]. Yellow and orange lines indicate the South Pole Queen Maud Land Traverse routes [*Picciotto et al.*, 1971]. The background shows elevation contour lines in 100 m steps. (b) Traverse route within the Antarctic continent.

[*Vaughan et al.*, 1999; *Giovinetto and Zwally*, 2000; *Arthern et al.*, 2006]. A comprehensive summary of ground-based field methods, their utilization and limitations is given by *Eisen et al.* [2008]. However, large parts of the vast East Antarctic plateau remain uncovered by ground-based measurements needed for these continent-wide interpolations, stressing the necessity of obtaining thorough data sets from this area. Even though coastal areas have a more important impact on current sea-level change [*Vaughan*, 2005] and are in places also poorly covered, accumulation data from sites far inland are needed in addition to coastal records to understand spatial and temporal accumulation patterns on the plateau areas and the outflow glaciers. The Norwegian-US IPY traverse through East Antarctica is dedicated to close some of these data gaps by contributing significant field data and assessing the spatiotemporal pattern of accumulation rates on remote areas of the East Antarctic plateau. The aim of the project is to improve understanding the mass balance of East Antarctica and its contribution to sea-level change, provide field data for calibration of model assessments as well as satellite-based estimates, and obtain information about climate signals and their changes within the last decades to about one millennium. Another important aspect of the Norwegian-US traverse project is to revisit sites measured during the South Pole Queen Maud Land Traverses (SPQMLT) [*Picciotto et al.*, 1971] and assess the legacy of these older data sets by updated records. Here, we report accumulation rates obtained from volcanic chronology based on dielectric profiling (DEP) of four shallow firn cores retrieved during the first leg of the traverse in the Antarctic summer season 2007/2008. Another firn core is analyzed for chemistry and we use the sulphur data to obtain accumulation rates from volcanic events. We discuss temporal variability within the cores as well as spatial variability around the drill sites. Ground-penetrating radar is utilized to address the spatial representativity of our firn cores. By comparing the conductivity records obtained from dielectric profiling with the sulphate records from chemistry analysis we also show the validity of the DEP method for dating where there are no chemistry data available.

2. Study Area

[4] The traverse route is depicted in Figure 1, going from Norwegian Troll Station through large parts of Dronning Maud Land (DML) to South Pole, covering about 2600 km in total. (Note that Dronning Maud Land and Queen Maud Land refer to the same geographic area. We use the official term Dronning Maud Land in the text, yet the SPQMLT traverses refer to it as Queen Maud Land which is cited here

Table 1. Depth of the Tambora/1809 Layers in Different Firn Cores Along the Traverse Line and in the Area[a]

Core Name	Year Drilled	Lat	Long	Elevation (m a.s.l.)	Depth Tambora (m)	Depth 1809 (m)	Acc Since Tambora (kg m^{-2} a^{-1})	Source
Site I	2007	73°43′S	7°59°E	3174	20.70	21.19	52	this paper
Site M	2001	75°00′S	15°00′E	3457	18.84	19.36	43	*Hofstede et al.* [2004]
NUS07-3	2007	77°00′S	26°03′E	3582	10.89	11.33	22	this paper
NUS07-4	2007	78°13′S	32°51′E	3595	10.33	10.64	19	this paper
NUS07-6	2008	80°47′S	44°51′E	3672	8.98	9.12	16	this paper
NUS07-8	2007	84°11′S	53°32′E	3452	13.57	13.86	32	this paper
Plateau Remote	1987	84°00′S	43°00′E	3330	15.10	15.66	40	*Cole-Dai et al.* [2000]
South Pole	2001	90°00′S	0°00′E	2850	30.44	31.37	76.5	*Budner and Cole-Dai* [2003], *Mosley-Thompson et al.* [1999]

[a]Long, longitude; Lat, latitude; Acc, accumulation. The value for South Pole is an average of six cores as reported by *Mosley-Thompson et al.* [1999].

accordingly where SPQMLT data are concerned.) Along the route more than 700 m of firn cores were drilled for a variety of purposes, including four shallow cores (19–25 m deep) for DEP analysis (NUS07-3, NUS07-4, NUS07-6, and NUS07-8, see Table 1 for coordinates of firn-core sites). Previous accumulation data from this region are sparse, although the more northerly parts of DML have been covered among others by Norwegian, Swedish, Japanese, and German expeditions [*Isaksson and Karlen*, 1994; *Isaksson et al.*, 1999; *Oerter et al.*, 1999; *Karlöf et al.*, 2000, 2005; *Winther et al.*, 2002; *Hofstede et al.*, 2004; *Rotschky et al.*, 2004; *Anschütz et al.*, 2007, 2008; *Eisen et al.*, 2005; *Takahashi et al.*, 1994]. However, especially the eastern part of DML and the interior of the remote plateau remain largely uncovered by these expeditions. In the 1960s US-led expeditions covered parts of the interior plateau under SPQMLT I–III [*Picciotto et al.*, 1971] (see Figure 1 for their traverse routes). They report accumulation values ranging between 6 and 70 kg m^{-2} a^{-1}. Some more recent studies from firn cores report an increase in 20th century accumulation in parts of the investigation area, including South Pole [*Oerter et al.*, 1999; *Mosley-Thompson et al.*, 1999], however, the trend is not significant in the study by *Oerter et al.* [1999]. This paper deals with accumulation rates over the last 200 years and contributes significant insight in accumulation variability in addition to those older data sets.

3. Methods

[5] Dielectric profiling (DEP) is a nondestructive technique for high-resolution measurements of electrical conductivity and dielectric permittivity. The device essentially consists of a guarded capacitor moving along the core [*Wilhelms*, 1996]. The details of the underlying physics [*Glen and Paren*, 1975] as well as the device [*Wilhelms et al.*, 1998; *Moore et al.*, 1989] are discussed elsewhere and will not be described here.

[6] We measured our firn cores with the DEP instrument at Norwegian Polar Institute (NPI) in the cold laboratory. Measurements were carried out in 5 mm intervals and the permittivity and conductivity were calculated from the conductance and capacitance, using the permittivity of solid ice and the free-air reading of the DEP device [*Wilhelms*, 1996]. In principle the temperature should also be accounted for, however, our measurements were carried out in the laboratory at a constant temperature of −20°C and the cores have been stored at that temperature before the measurements long enough to adjust. Hence, a temperature correction is not mandatory [*Wilhelms*, 1996]. In order to compare conductivity peaks, the conductivity record has to be corrected for density fluctuations [*Wilhelms*, 1996; *Glen and Paren*, 1975]. We used the complex Looyenga mixing model [*Looyenga*, 1965] to calculate density from permittivity, which was then used for correction of the conductivity signal. The procedure is outlined by *Glen and Paren* [1975] and *Wilhelms* [1996] and links the high-frequency conductivity to the conductivity of solid ice and the volume ratio (essentially the density of the firn divided by the density of solid ice). As conductivity and density of solid ice are known (19 μS m^{-1} and 918 kg m^{-3}, respectively [*Wilhelms*, 1996]) the density can be calculated from the measured conductivity. The absolute of the Looyenga-based density varies from 179 to 663 kg m^{-3} in our DEP cores (Figure 2), similar to the values reported by other studies from the East Antarctic plateau [*Rotschky et al.*, 2004; *Hori et al.*, 1999]. The largest variations occur in the upper few meters where the core quality is generally lower. In addition to the Looyenga-based density, bulk density was obtained in the field, using the weight of the ice core pieces measured with an electronic balance. The core diameter was measured manually using a caliper. In comparison with the Looyenga-based density the bulk density shows slightly higher values

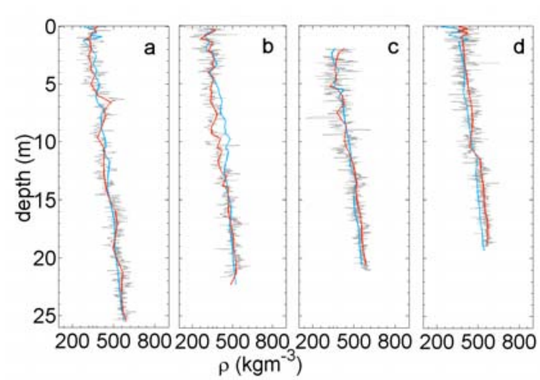

Figure 2. Comparison of density calculated according to the complex Looyenga mixing model (dark grey line) and bulk density from diameter, length, and weight of individual core sections (blue line). For comparison the Looyenga-based density averaged over the lengths of the individual bulk pieces is also plotted (red line). (a) NUS07-3, (b) NUS07-4, (c) NUS07-6, and (d) NUS07-8.

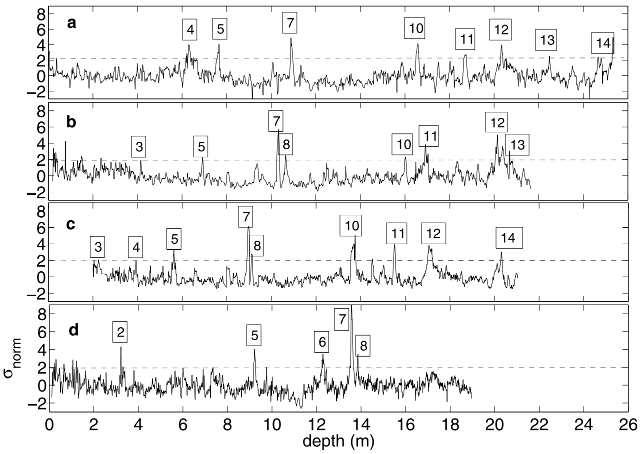

Figure 3. Processed normalized conductivity records from the four DEP cores. (a) NUS07-3, (b) NUS07-4, (c) NUS07-6, and (d) NUS07-8. The dashed line indicates the twofold standard deviation $2\sigma_{norm}$. Volcanic peaks are indicated by numbers; the names of the respective volcanoes and the ages of layers are given in Table 2.

in some of the shallow sections, possibly due to varying core diameter (Figure 2). The average difference between bulk density and Looyenga-based density is 3.5%. However, *Hawley et al.* [2008] show that errors in density have a significantly lower effect on errors in accumulation rates than the uncertainty in actual layer depth. Thus, we assume that the difference between the density calculations does not introduce larger errors in our final accumulation results (see section 4.1 for a discussion of general error sources and the average accumulation error).

[7] In order to obtain accumulation rates, a depth-age scale based on volcanic chronology has been established from the conductivity records by comparison with known records [*Hofstede et al.*, 2004; *Traufetter et al.*, 2004]. Detection of volcanic peaks in the conductivity signal follows the procedures outlined by *Hofstede et al.* [2004] and other authors, i.e., it is assumed that a peak above the twofold standard deviation ($2\sigma_{norm}$ in Figure 3) is related to a volcanic eruption. Thus, we normalized the conductivity records by subtracting the mean and dividing by the standard deviation (Figure 3). However, not all peaks could be assigned to known volcanoes and especially the upper 2 m exhibit rather noisy behavior so they were left out. The top 2 m of core NUS07-6 were not analyzed at all because of very poor core quality. From the depth of the dated horizon the accumulation can be calculated using the age of the layer and information about the density. Thus, mean accumulation rates over the respective time periods determined by the dates of the different eruptions have been obtained (Table 2).

[8] The 30 m long firn core from site I (Figure 1) was analyzed at Desert Research Institute, Reno, USA, using a sophisticated combination of continuous flow analysis and inductively coupled plasma mass spectrometry (ICP-MS) [*McConnell et al.*, 2002]. The core from site M (Figure 1) which we base our dating on has been retrieved in 2001 during a Norwegian expedition [*Winther et al.*, 2002] and was analyzed at NPI for DEP and at Alfred Wegener Institute, Bremerhaven, Germany, for chemistry. The DEP data and depth-age scale have been discussed by *Hofstede et al.* [2004].

[9] Ground-penetrating radar (GPR) is used to investigate firn stratigraphy, accumulation rate and its spatial patterns [*Richardson et al.*, 1997; *Richardson and Holmlund*, 1999; *Eisen et al.*, 2005, 2008; *Pälli et al.*, 2002]. GPR profiles were collected along the entire traverse route (Figure 1). The system used is a frequency-modulated continuous-wave (FM CW) radar [*Hamran and Langley*, 2006] with a center frequency of 5.3 GHz and a bandwidth of 1 GHz. The vertical resolution in firn is approximately 0.1 m (depending on the dielectric properties of the firn pack) and the achieved penetration depth is 20 m assuming a mean firn-pack velocity of 0.23 m ns^{-1}. In order to identify a specific layer from the DEP measurements in the GPR profile the

Table 2. Volcanic Peaks and Accumulation Rates With Absolute Errors for Cores NUS07-3, NUS07-4, NUS07-6, and NUS07-8 and Site I[a]

			NUS07-3		NUS07-4		NUS07-6		NUS07-8		Site I	
Peak Number	Volcano	Year AD	Acc	Error	Acc	Error	Acc	Error	Acc	Error	Acc	Error
1	Pinatubo[b]	1991	-	-	-	-	-	-	-	-	52	6.4
2	Agung[b,c]	1963	-	-	-	-	-	-	30	2.7	56	4.7
3	Cerro Azul[b]	1932	-	-	19	1	12	0.4	-	-	-	-
4	Santa Maria[d]	1902	21	0.8	-	-	14	0.3	-	-	-	-
5	Krakatau[b,c]	1883	23	0.8	19	0.6	13	0.4	32	1.1	53	1.9
6	Coseguina[b,c]	1835	23	0.6	19	0.4	14	0.3	31	0.8	52	1.5
7	Tambora[b,c]	1815	22	0.5	19	0.4	16	0.3	32	0.7	52	1.3
8	unknown[b,c]	1809	22	0.5	19	0.4	15	0.3	32	0.7	52	1.3
9	Peteroa[b,c]	1762	-	-	-	-	-	-	-	-	52	1.1
10	unknown[c,e]	1695	22	0.4	19	0.3	16	0.3	-	-	-	-
11	Gamkonora[b,c]	1673	24	0.4	20	0.3	18	0.3	-	-	-	-
12	Deception Island[b,c]	1641	24	0.4	22	0.3	20	0.3	-	-	-	-
13	unknown[c,e]	1622	26	0.4	21	0.3	-	-	-	-	-	-
14	Huayanaputina[b,c]	1600	28	0.4	-	-	21	0.3	-	-	-	-

[a]Accumulation rates are between year of eruption and present. All values are given in kg m^{-2} a^{-1}.
[b]*Traufetter et al.* [2004].
[c]*Hofstede et al.* [2004].
[d]*Moore et al.* [1991].
[e]*Cole-Dai et al.* [2000].

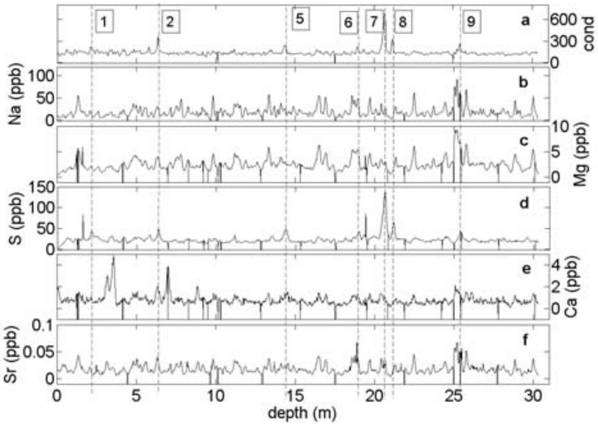

Figure 4. Chemistry data from site I core. Volcanic peaks are indicated by numbers; see Table 2. The sudden drops indicate that there are no data available for these depths (represented by an arbitrary value of -1, shown here for the sake of completeness). (a) Liquid conductivity, (b) sodium, (c) magnesium, (d) sulphur, (e) calcium, and (f) strontium.

mean radar-wave velocity v_{med} down to the layer depth was calculated using the speed of light in vacuum c_0 and the average permittivity value ϵ:

$$v_{med} = \frac{c_0}{\sqrt{\epsilon}}$$

4. Results and Discussion

4.1. Dating the Firn Cores

[10] Several volcanoes could be identified in the DEP records, the most obvious one being the double peak of the eruption of Tambora in 1815 and an unknown volcano in 1809 (Figure 3). These peaks have been observed widely throughout Antarctica [*Legrand and Delmas*, 1987; *Langway et al.*, 1995; *Frezzotti et al.*, 2005; *Hofstede et al.*, 2004; *Karlöf et al.*, 2000; *Isaksson et al.*, 1999; *Oerter et al.*, 1999; *Cole-Dai et al.*, 2000; *Traufetter et al.*, 2004] and are used as time markers here. Based on that, it was possible to assign the eruptions of Krakatau (1883) [*Hofstede et al.*, 2004; *Traufetter et al.*, 2004; *Oerter et al.*, 1999] for all of the cores and further eruptions like the one from Deception Island of 1641 [*Traufetter et al.*, 2004; *Hofstede et al.*, 2004] for three of the four cores (Figure 3). Table 2 gives an overview of identified eruptions and their respective ages within the four DEP cores as well as accumulation rates averaged over the time period from the eruption to present (2007).

[11] The sulphur data from site I were used to establish a volcanic chronology as there are no DEP data available for this core. Detected volcanoes and accumulation rates are given in Table 2. As in the DEP cores the Tambora/1809 double peak serves as time marker, enabling us to assign the eruptions of Agung (1963), Krakatau (1883) and Coseguina (1835) [*Hofstede et al.*, 2004; *Traufetter et al.*, 2004]. In order to confirm the volcanic origin of the conductivity peaks, we compare liquid conductivity, sulphur and sea salt related chemical species from the site I core with our DEP records since DEP responds to both acidity and salinity [*Moore et al.*, 1989, 1991; *Karlöf et al.*, 2000]. Figure 4 shows the profiles of these species for the entire depth of the site I core, highlighting the prominent peaks. It is clearly visible that e.g. the Tambora/1809 double peak corresponds very well in the conductivity and sulphur records whereas the sea salts do not show such prominent peaks at these specific depths. Thus, it can be concluded that those conductivity peaks originate from enhanced sulphur and not from enhanced sea salt concentrations. The same holds for the core from site M (Figure 1), where conductivity from DEP and sulphate curves compare well, both clearly showing the eruptions of Tambora 1815 and unknown 1809 (Figure 5). For comparison, we also plot chloride (the only sea salt record available for this core) which does not peak at the Tambora/1809 depths. *Hofstede et al.* [2004] based their analysis of the site M core on DEP-derived conductivity. Our Figure 5 indicates that their conductivity peaks coincide with sulphate peaks, thus showing the volcanic origin and the validity of their time scale for the site M core. However, some of the conductivity peaks coincide not only with enhanced sulphur/sulphate but also with enhanced sea salts in Figures 4 and 5, making the origin of the respective conductivity peaks unclear. We thus focus in our discussion mainly on the peaks from Krakatau, Tambora/1809 and Deception Island which do not coincide with enhanced sea salts. Therefore we assume that these peaks are of volcanic origin in the site M core and in our DEP cores and provide a reliable source for dating the cores.

[12] Uncertainties in our accumulation results arise mainly from dating which is based on the Tambora/1809 double peak as absolute time marker. Annual-layer counting was not feasible due to the very low accumulation rates. Thus, we have to rely on time markers and peak matching by comparison with well-dated cores from the wider area as for example the core from site M [*Hofstede et al.*, 2004]. Dating uncertainty of core M is 3 years [*Hofstede et al.*, 2004] which is similar to the time scale from other East Antarctic

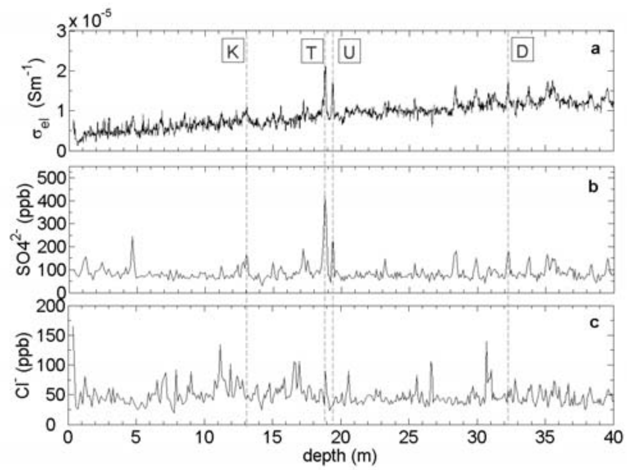

Figure 5. DEP and chemistry records from site M. Only the upper 40 m are shown here. (a) Conductivity from DEP, (b) sulphate, and (c) chloride. Vertical lines and letters mark the eruptions of Krakatau (K), Tambora (T), unknown 1809 (U), and Deception Island (D) as reported by *Hofstede et al.* [2004].

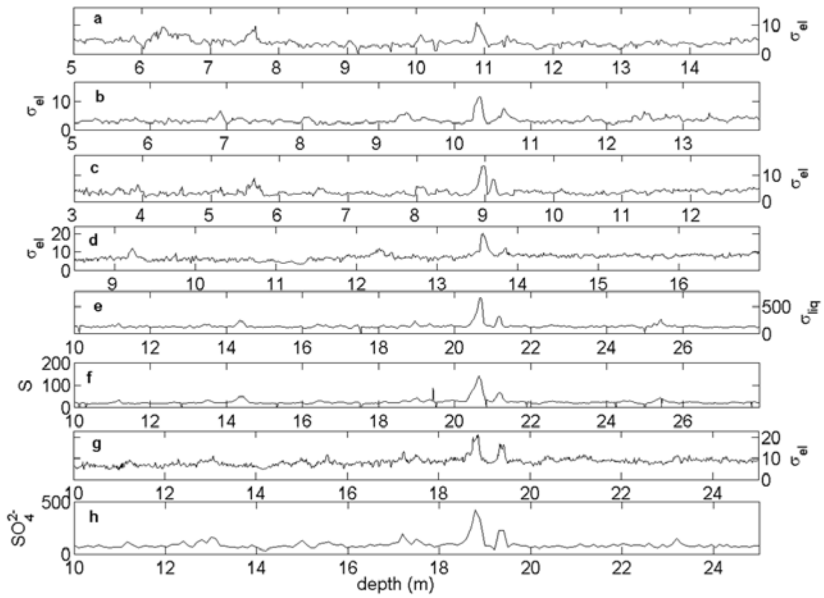

Figure 6. Comparison of the shape of the Tambora/1809 double peak in different cores. (a) Conductivity for core NUS07-3, (b) conductivity for core NUS07-4, (c) conductivity for core NUS07-6, (d) conductivity for core NUS07-8, (e) liquid conductivity for the site I core, (f) sulphate site I, (g) conductivity for the core from site M, and (h) sulphur for core M. (Conductivity values σ_{el} are given in μS m^{-1}; S and SO_4^{2-} are given in ppb.)

cores dated by chemistry or DEP [*Traufetter et al.*, 2004]. *Hofstede et al.* [2004] and *Traufetter et al.* [2004] assume that this error is constant back to 1600 AD and only increases for time spans longer than that which is beyond the scope of our study. Thus, it is reasonable to assume a constant uncertainty of 3 years also for our cores. Arguably uncertainty would be higher once annual accumulation is considered but as we base our results only on the volcanic chronology it is reasonable to assume a constant uncertainty for the dated strata in accordance with *Traufetter et al.* [2004] and *Hofstede et al.* [2004]. Uncertainty in the density estimates is linked to accuracy of the DEP measurements which is about 1% [*Wilhelms*, 1996]. This leads to an uncertainty of approximately 1% in the density, derived from error propagation. Using a depth uncertainty of 2 cm resulting from the placement of the core on the DEP bench and minor loss of core material during transport and the above mentioned uncertainties of density and time, we are able to estimate the overall accumulation uncertainty by error propagation. Thus, relative errors for accumulation are in the range of 1.4 to 21% with the largest errors for the shorter time scales. The average relative error is about 8% and the error for accumulation rates averaged over the period 1815–2007 is around 2.3% which is comparable to the values reported by *Frezzotti et al.* [2005, 2007]. The absolute errors are in the range of 0.3 to 6.4 kg m^{-2} a^{-1} (Table 2). Note that absolute errors are higher for site I in Table 2 but accumulation at site I is higher than at the other sites and thus relative errors are similar for all our core sites.

[13] Even though we do not have sulphur and sea salt records available for our DEP cores, the shape of the double peak of the Tambora/1809 layers (Figure 6) enables us to date these cores with confidence. Over all, the firn cores along the traverse line (site I core, site M core, NUS07-3, NUS07-4, NUS07-6, NUS07-8) show a high similarity in the shape of these peaks, either in the sulphur/sulphate records or in the conductivity data (Figure 6). This similarity between sulphate/sulphur records and conductivity records allows us to safely assume volcanic origin for our conductivity peaks and base our dating upon that assumption. It should be noted though that generally non sea salt (nss) sulphate is used in literature [*Legrand and Delmas*, 1987; *Langway et al.*, 1995] which is derived from the sodium record and the ratio of sulphate to sodium in seawater [*Traufetter et al.*, 2004]. However, for site I we do not have sulphate but sulphur and for site M there is no sodium available. Thus, we use the sulphur (site I) and total sulphate (site M) records here. Since we are only concerned with peak detection and not with a quantitative determination of volcanic fluxes these records should be sufficient [*Traufetter et al.*, 2004]. The comparison of sea salt records, sulphur/sulphate data and conductivity (Figure 6) confirms that DEP-based conductivity can be used for dating for firn cores where there are no chemistry data available and that DEP provides a valuable tool for quickly determining accumulation rates.

4.2. Temporal Variability of Accumulation Rates

[14] The double peak of Tambora 1815 and the unknown volcano of 1809 provides a clear time marker for dating. Using these peaks as reference horizons, we are able to obtain accumulation rates averaged over the last ≈200 years. Over all, the accumulation rates have been fairly stable since 1815 (Table 2). The core from site I does not show any significant change in accumulation since the Tambora eruption. Site NUS07-3 shows a slight increase for 1883–2007 (Krakatau to present) in relation to 1815–2007. At site NUS07-4 no change between these two periods occurs,

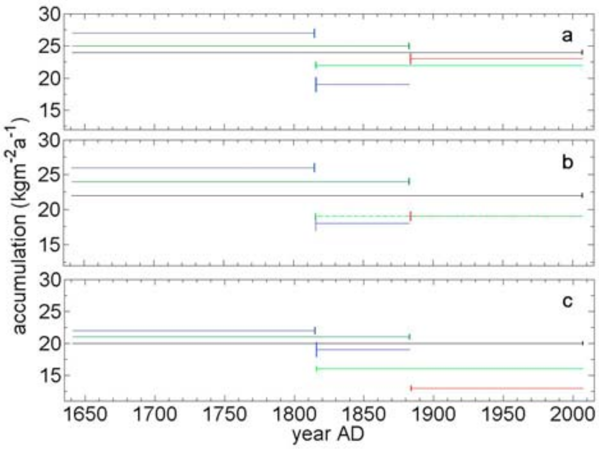

Figure 7. Accumulation rates for the three DEP cores that reach farther back than Tambora 1815. Selected time periods shown here are: 1883–2007 (red line), 1815–2007 (bright green line), 1641–2007 (black line), 1815–1883 (bright blue line), 1641–1883 (dark green line), and 1641–1815 (dark blue line). The vertical bars show the respective errors; their position on the *x* axis is chosen arbitrarily for better readability, but error values are representative for the entire time period depicted. (a) NUS07-3, (b) NUS07-4, and (c) NUS07-6.

whereas sites NUS07-6 and NUS07-8 show a slight decrease in 1883–2007 relative to 1815–2007 (Table 2 and Figure 7). The accumulation rate for the period 1815–1883 (between the eruptions of Tambora and Krakatau) is slightly lower than the average over 1815–2007 at sites NUS07-3 and NUS07-4, slightly higher at site NUS07-6 and seems unchanged at site NUS07-8 (Figure 7). *Hofstede et al.* [2004] report a period of relatively higher accumulation rates for the early 19th century from six firn cores, including the site M core. This cannot be confirmed by our results (Figure 7 and Table 2). Along with some other authors [*Oerter et al.*, 1999; *Mosley-Thompson et al.*, 1999], *Hofstede et al.* [2004] also find increasing accumulation rates in the late 20th century. We do not have sufficient resolution for that, but some of our cores exhibit peaks that could be assigned to several more recent eruptions. Although these peaks are weak compared to the prominent Tambora/1809 peaks and have not been reported as widely in Antarctic ice cores, they give some information about the accumulation pattern in the 20th century. Site NUS07-3 and NUS07-6 seem to show the Santa Maria eruption of 1902 [*Moore et al.*, 1991], site NUS07-4 and NUS07-6 the one from Cerro Azul in 1932 [*Traufetter et al.*, 2004] and site NUS07-8 the eruption of Agung in 1963 [*Traufetter et al.*, 2004; *Hofstede et al.*, 2004; *Oerter et al.*, 1999]. While it is not possible to compare the exact time periods due to the different volcanic events observed in the different cores, these more recent eruptions enable us at least to determine accumulation rates over the 20th or late 20th century, respectively. All three cores show comparatively lower accumulation rates for the time period between the respective eruption and present (Table 2). Hence, accumulation increase for the (late) 20th century cannot be confirmed from our firn-core locations on the East Antarctic plateau.

[15] Only three of our four DEP cores (NUS07-3, NUS07-4, NUS07-6) allow us to access accumulation before the Tambora/1809 eruptions. Figure 7 shows accumulation averaged over the time periods between some of the volcanic horizons seen in these cores. Concerning temporal variability, there is a generally decreasing trend visible for all three sites over the 19th and 20th century. In order to quantify that decrease, we compare accumulation for the period 1641–1815 (i.e., between the eruptions of Deception Island and Tambora) with accumulation rates averaged over the time span 1815–2007 (Tambora to present). We chose these eruptions because the Tambora peak is a clear and absolute time marker and the Deception Island eruption is the earliest one visible in all three cores considered here. For the period 1815–2007 accumulation rates appear to have decreased by 24% at site NUS07-3, by 27% at site NUS07-4 and by 28% at site NUS07-6 in comparison to 1641–1815. Thus, these three cores seem to show a consistent trend of decreasing accumulation in the 19th and 20th century compared to at least ≈170 years before Tambora.

[16] Some other studies suggest an accumulation increase in Dronning Maud Land for the 20th or late 20th century [*Mosley-Thompson et al.*, 1999; *Oerter et al.*, 1999; *Hofstede et al.*, 2004]. However, in our data a recent trend by the end of the 20th century might well be averaged out when calculating accumulation from 1815 to present. For South Pole, *Mosley-Thompson et al.* [1999] report a long-term accumulation value of 76.5 kg m^{-2} a^{-1} (Tambora to present) and an increase up to 85 kg m^{-2} a^{-1} since the 1960s. As these values were obtained mostly upwind of South Pole Station, they should not be disturbed by the station building but reflect a real increase in accumulation [*Mosley-Thompson et al.*, 1999]. At Dome C *Frezzotti et al.* [2005] find a recent accumulation increase to 32 kg m^{-2} a^{-1} in comparison with the long-term mean of 25.3 (1815–1998). They also report a general accumulation increase for several drill sites along the transect from Terra Nova Bay to Dome C. In their cores, the period 1966–1998 shows a 14 to 55% higher accumulation than the period 1815–1998. *Karlöf et al.* [2000] find a decrease of 8% at Camp Victoria for the period 1452–1641 in comparison with the long-term mean (1259–1997). They also report a slight decrease for the time period 1933–1970. *Stenni et al.* [2002] find that accumulation at Talos Dome had increased by 11% over the 20th century in comparison to the 800 year mean. However, *Frezzotti et al.* [2007] report no significant increase in accumulation over the last two centuries near Talos Dome. Yet *Urbini et al.* [2008] find significantly lower accumulation rates in the Southwest of Talos Dome during 1835–1920 with respect to 1920–2001.

[17] *Isaksson et al.* [1996] observe an accumulation decrease over the period 1932–1991 from a coastal core in Dronning Maud Land and report no trend in accumulation for the period 1865–1991 from another core on the plateau (75°S, 2°E, 2900 m asl). Moreover, *Isaksson et al.* [1999] find that a general accumulation increase in especially the latter part of the 20th century is not necessarily valid for the whole polar plateau of Dronning Maud Land. They discuss temporal variability within their different firn cores (C–M; see Figure 1) between the periods 1955–1965 and 1965–1996 and conclude that accumulation has been decreasing at the higher-elevation parts of the plateau

(above 3250 m). For sites C–H though accumulation is reported to have increased by 15–50% during the latter period, with the exception of site F (Figure 1). *Isaksson et al.* [1999] conclude that those sites are possibly more affected by cyclonic activity since they are closer to the coast than the high-elevation sites I–M. The accumulation decrease at these latter sites is in agreement with our results, although the accumulation rates observed in our study arguably are averaged over a longer time period and thus neglect (multi)annual or decadal variability. Furthermore, *Genthon et al.* [2009] show that accumulation in Eastern Wilkes Land has been lower than previously assumed and *Magand et al.* [2007] demonstrate that older data sets are often biased and tend to overestimate accumulation on the polar plateau.

[18] The different studies and conclusions highlight the fact that the temporal accumulation pattern on the polar plateau is complicated and not well understood. Comparison between the individual studies is difficult due to different observation periods and poor spatial coverage. Our analysis provides new and reliable accumulation data from yet uncovered areas and insight in temporal accumulation variability at the drill sites that fit the results of *Genthon et al.* [2009], *Magand et al.* [2007], *Isaksson et al.* [1999] and other authors.

4.3. Spatial Variability

4.3.1. General Accumulation Distribution

[19] Mean accumulation rates from the four traverse cores range from 16 to 32 kg m^{-2} a^{-1} averaged over the last ~200 years (Tambora to present). There is a sightly decreasing trend from site NUS07-3 to site NUS07-6, whereas site NUS07-8 shows comparatively higher accumulation. The most plausible explanation for this is increasing continentality, increasing elevation (from site NUS07-3 to NUS07-6) and lower elevation for site NUS07-8 (see Table 1 for elevation of the individual drill sites). Site NUS07-8 is also the closest one to South Pole where *Mosley-Thompson et al.* [1999] report an average accumulation rate of 76.5 kg m^{-2} a^{-1} since 1815. Even though the accumulation at South Pole is arguably much higher than our results a spatial increase of accumulation towards site NUS07-8 as shown in our data fits the expected pattern well.

[20] Table 1 gives an overview of the depth of the Tambora/1809 layers in different firn cores from this area of the East Antarctic plateau and deduced accumulation rates over approximately the last 200 years (between 1815 and the drilling date of the respective core). While the distances between drill sites are large and thus inhibit a detailed discussion of spatial variability or interpolation between individual sites, a general pattern becomes obvious nevertheless. Not surprisingly, accumulation generally decreases with increasing elevation and with increasing continentality. Our results are on the lower edge of the range of accumulation values (Table 2), yet they fit in the general picture and show that accumulation, when averaged over two centuries, is slightly lower than expected from previous data from the area [e.g. *Picciotto et al.*, 1971].

[21] In general comparison with other data is difficult due to the large spatial distances. Thus, we only give a brief review of SPQMLT and other data from the area around the traverse route and discuss the general accumulation pattern. When assessing older data it is also important to keep quality control in mind as explained by *Magand et al.* [2007] who show that those older data are likely to introduce biases. In line with *Genthon et al.* [2009], *Magand et al.* [2007] suggest careful data selection and exclude accumulation obtained by stratigraphy alone from their studies. This would apply to some of the SPQMLT data [*Picciotto et al.*, 1971]. However, as these provide the largest data set available for our investigation area, they are cited here nevertheless stressing that the comparison can only be viewed with caution. *Picciotto et al.* [1971] report accumulation rates of some 27 kg m^{-2} a^{-1} for Plateau Station and 31 kg m^{-2} a^{-1} for Pole of Relative Inaccessibility for the period 1955–1965. Their accumulation rates along the three stretches of the SPQMLT are in the range of 6 to 70 kg m^{-2} a^{-1} where the very low accumulation zones (6 to 10 kg m^{-2} a^{-1}) are concentrated around the region 81°–82°S, 20°E. By comparison, our accumulation rates are on the lower edge of their values. However, spatial distance limits the comparison and the observation time periods are not the same, as *Picciotto et al.* [1971] obtained their values mainly from radioactive layers in 1955 and 1965 and from snow-pit stratigraphy, thus covering shorter time spans. Our accumulation rates, on the other hand, are averaged over 200 years or more and hence do not account for shorter scale (decadal to annual) variability. *Cameron et al.* [1968] report accumulation values from stakes at Pole of Relative Inaccessibility. Their value averaged over six years is 36 kg m^{-2} a^{-1}, slightly higher than the values reported by *Picciotto et al.* [1971]. Our sites NUS07-6 (16 kg m^{-2} a^{-1}) and NUS07-8 (32 kg m^{-2} a^{-1}) exhibit comparatively lower accumulation rates, yet both sites are about 220 km away from the Pole of Relative Inaccessibility and thus a direct comparison is difficult. At Plateau Station accumulation data have been obtained from snow-pit stratigraphy of 64 shallow snow pits and a stake farm [*Koerner*, 1971]. The overall average over a period of 10–12 years is reported as 28 ± 4 kg m^{-2} a^{-1} [*Koerner*, 1971], agreeing well with the reports from *Picciotto et al.* [1971]. Our sites NUS07-4 (19 kg m^{-2} a^{-1}) and NUS07-6 (16 kg m^{-2} a^{-1}) are about 200 km away from Plateau Station so a direct comparison is again limited by distance.

[22] *Mosley-Thompson* [1996] and *Cole-Dai et al.* [2000] report accumulation values from Plateau Remote (84°S, 43°E, 3330 m asl.) of around 40 kg m^{-2} a^{-1}. Dome Fuji is reported to have an average value of 27 kg m^{-2} a^{-1} (1995–2006) [*Kameda et al.*, 2008], which is more in the range of our values. Although the area is poorly covered, our results fit in the general picture of very low accumulation on the East Antarctic plateau. *Frezzotti et al.* [2005] report accumulation data along a traverse from Terra Nova Bay to Dome C. Their firn-core based results are in the range of 17 to 98 kg m^{-2} a^{-1} for the period 1815–1998. Dome C itself has an accumulation of 25.3 ± 1 kg m^{-2} a^{-1} [*Frezzotti et al.*, 2005]. Even though the distances to our study area are large, this shows that the general pattern of low accumulation for the high elevation sites is maintained over much of the East Antarctic plateau.

[23] *Isaksson et al.* [1999] report accumulation data from radioactive layers in 1955 and 1965 observed in several shallow firn cores along a traverse line up to site M (Figure 1), reporting values between 23 to 123 kg m^{-2} a^{-1}. Not

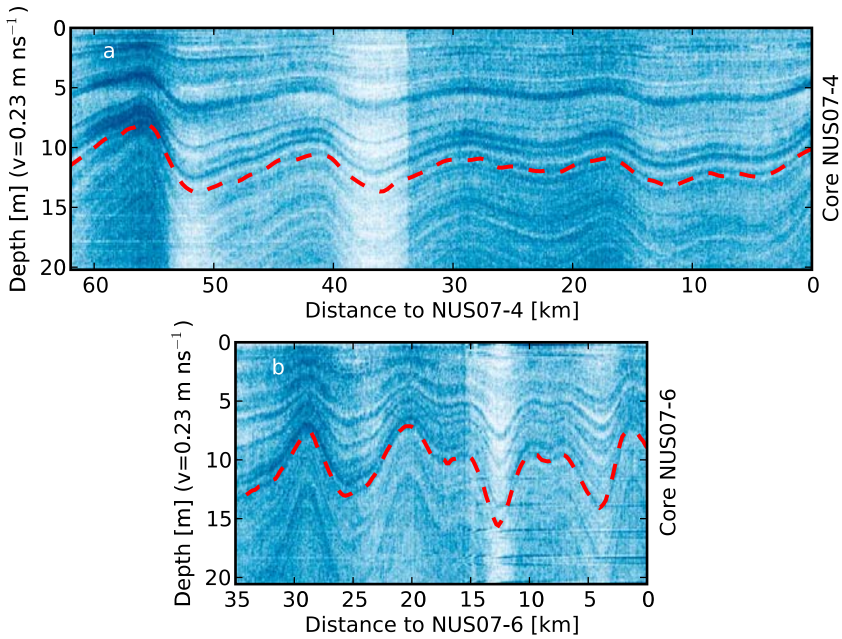

Figure 8. The 5.3 GHz GPR profiles leading southward to drill sites (a) NUS07-4 and (b) NUS07-6. The presumed Tambora layer is indicated by a dashed red line in both profiles. The average accumulation in Figure 8a is 21 kg m^{-2} a^{-1} and in Figure 8b is 19 kg m^{-2} a^{-1}. Both profiles represent low-accumulation areas exhibiting only gentle spatial variations in accumulation rates.

surprisingly, their accumulation data basically decrease with increasing distance from the coast and increasing elevation. This pattern agrees well with our findings. For site I, *Isaksson et al.* [1999] report an accumulation rate of 52 ± 4 kg m^{-2} a^{-1}, well within the range of our values for this site (Table 2). Hence, our analysis confirms that the accumulation pattern at site I has been fairly stable not only since 1955 but also over the last 200 years.

4.3.2. Spatial Representativity From GPR Data

[24] GPR has been used widely to assess the question of how representative single firn cores are for the area around them [*Richardson and Holmlund*, 1999; *Frezzotti et al.*, 2004, 2005; *Pälli et al.*, 2002]. While we cannot fully quantify the spatial representativity of our firn cores yet (GPR data and shape of radar layers will be discussed in detail in forthcoming papers [e.g., *Müller et al.*, 2009]) we use short sections of our traverse GPR recordings at NUS07-4 and NUS07-6 and earlier measurements by *Richardson and Holmlund* [1999] between sites I and M to get a glimpse on the spatial representativity of the cores. In our firn core from site NUS07-4 the Tambora eruption was identified at a depth of 10.33 m (Table 1). A layer at that depth at NUS07-4 is traceable in the GPR profile for 62 km northwards from the core (Figure 8a). Due to temperature sensitivity of the GPR system there are gaps in the data farther north prohibiting farther tracing. Assuming laterally constant firn density along the profile the mean accumulation rate over the 62 km is 21 kg m^{-2} a^{-1} with a standard deviation of 2.5 kg m^{-2} a^{-1} or 12% from the mean value. The minimum and maximum values are 14.4 and 25 kg m^{-2} a^{-1}, respectively, deviating by 30% and 19% from the mean. The accumulation rate from the core NUS07-4 (19 kg m^{-2} a^{-1}) deviates by 9% from the mean accumulation along the GPR profile, thus well within the standard deviation for this stretch, along which there is also a northward trend toward higher accumulation rates.

[25] The Tambora event was found at a depth of 8.98 m in the NUS07-6 core, yielding an accumulation rate of 16 kg m^{-2} a^{-1} (Table 1). In the GPR profile coming from the North to the core site, the Tambora layer is traceable over a distance of 35 km (Figure 8b). The mean accumulation for the time period 1815–2007 along the 35 km GPR profile is 19 kg m^{-2} a^{-1} with a standard deviation of 3.9 kg m^{-2} a^{-1} or 19%. The minimum value is 13 kg m^{-2} a^{-1} or 33% of the mean value and the maximum was found to be 28 kg m^{-2} a^{-1}, deviating by 46% from the mean. The accumulation rate from the firn core NUS07-6 is 16 kg m^{-2} a^{-1}, about 15% less than the mean accumulation along the GPR profile and again well within the standard deviation of accumulation variability along the 35 km profile.

[26] *Richardson and Holmlund* [1999] showed by analysis of radar-layer depth and shape that the accumulation pattern from internal GPR horizons between sites I and M (Figure 1) is comparatively smooth and does not show large spatial variations around these sites. The standard deviation of layer depth around their drill sites is 2 to 31% where maximum and minimum depth deviate by up to 75% from the mean value. They conclude from their analysis that the representativity of their firn cores is generally high. The trend of decreasing variability with increasing elevation and increasing continentality is in agreement with the values seen around NUS07-4 and 6. Thus, the variability of GPR layers around our drill sites is well within the range of variability reported by *Richardson and Holmlund* [1999].

[27] *Frezzotti et al.* [2004, 2005] report spatial variability from radar layers around their drill sites along a traverse from Terra Nova Bay to Dome C. The standard deviation of their GPR-based accumulation ranges from 3 to 47% and

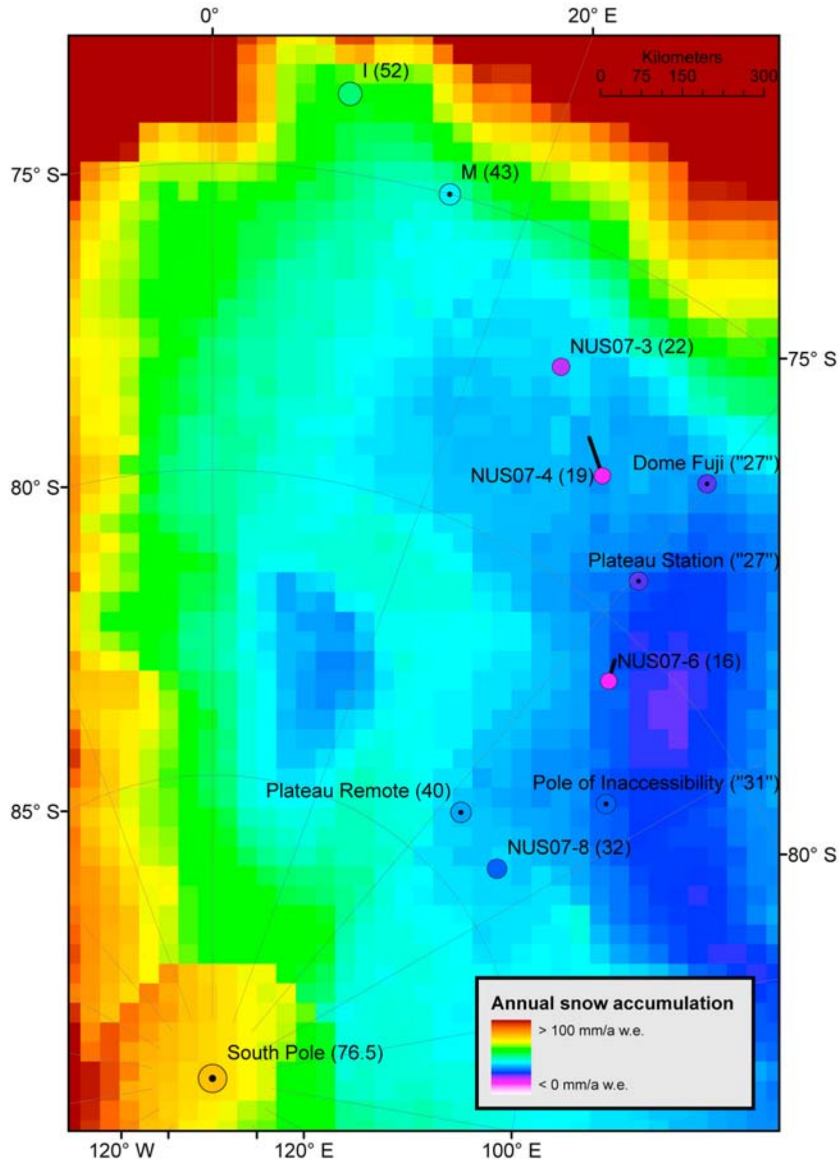

Figure 9. Accumulation map for the area around the traverse route based on the compilation by *Arthern et al.* [2006]. Circles indicate the locations of firn cores where numbers in parentheses give the accumulation rate for the period Tambora to present. Values at Plateau Station and Pole of Relative Inaccessibility have been taken from *Picciotto et al.* [1971] and are given in quotation marks as they do not refer to the period Tambora to present. The same holds for the value at Dome Fuji obtained from *Kameda et al.* [2008]. Circles with dots show core sites from external studies relevant for this study [*Picciotto et al.*, 1971; *Hofstede et al.*, 2004; *Cole-Dai et al.*, 2000; *Mosley-Thompson et al.*, 1999; *Kameda et al.*, 2008]. See Table 1 for details. The two black lines at sites NUS07-4 and NUS07-6 indicate the GPR profiles shown in Figure 8. The background map is obtained from *Arthern et al.* [2006].

the respective minima and maxima deviate by 12 to 85%, comparable to the results from *Urbini et al.* [2008] for spatial variability of GPR layers around Talos Dome. Thus, the spatial variability along our GPR transects agrees with the range reported by *Frezzotti et al.* [2004, 2005] and *Urbini et al.* [2008]. We therefore assume that our cores are well representative for the area around them and yield reliable accumulation information for the period 1815–2007 along the parts of the traverse route reported here. However, with a general variation of GPR-based accumulation in the range of 10–20% over tens of kilometers comparison with SPQMLT data from Plateau Station and Pole of Relative Inaccessibility is limited.

4.3.3. Comparison With Large-Scale Accumulation Trends

[28] As our results provide insight in accumulation variability of a largely uncovered area, they can also serve for validation of reconstructions of accumulation time series

from e.g. precipitation fields. Recently, *Monaghan et al.* [2006] undertook such a study, using ERA-40 precipitation fields and ice core data to derive a continent-wide accumulation pattern. They found no significant long term change of accumulation rates over the period 1957–2005. If at all, accumulation had been decreasing over parts of the East Antarctic ice sheet and our investigation area in the period 1995–2005 [*Monaghan et al.*, 2006, Figure 2F]. This is in agreement with our findings, although we cannot sufficiently resolve variability on the scale of decades. *Monaghan et al.* [2006] report values ranging between 20 and 50 kg m^{-2} a^{-1} for our study area. Thus, our values are largely on the lower edge of their large-scale pattern. Moreover, *Monaghan et al.* [2006] suggest that extending a study similar to theirs as far back as one to two centuries would help understanding climatic changes in Antarctica better. In this context our firn-core based accumulation data will be of valuable input to similar studies with a focus on longer-term time series, on the scale of up to two centuries. In addition, our firn cores are from an area that does not provide any records used by, e.g., *Monaghan et al.* [2006, Figure 1] for their assessment of continent-wide accumulation trends.

[29] *Arthern et al.* [2006] derive an accumulation map of Antarctica from a compilation of observations guided by satellite data from AMSR and AVHRR (Advanced Microwave Scanning Radiometer and Advanced Very High Resolution Radiometer, respectively) with a spatial resolution of ≈100 km. In comparison with the study by *Arthern et al.* [2006] our accumulation values at sites I and M (Table 2) fit quite well as *Arthern et al.* [2006] obtain 60 kg m^{-2} a^{-1} around site I and 48 kg m^{-2} a^{-1} around site M (Figure 9). However, for the sites NUS07-3, NUS07-4, NUS07-6 and NUS07-8 our values (Table 2) are clearly lower than the results by *Arthern et al.* [2006] who report 40 kg m^{-2} a^{-1} around NUS07-3, 36 kg m^{-2} a^{-1} around NUS07-4, 32 kg m^{-2} a^{-1} around NUS07-6 and 40 kg m^{-2} a^{-1} around NUS07-8 (Figure 9). Yet *Arthern et al.* [2006] obtain a spatial pattern similar to our results (Figure 9) with a decrease of accumulation from the area around site NUS07-3 to site NUS07-6, then an increase around site NUS07-8. As *Arthern et al.* [2006] used in situ observations from sites I and M for their study (their Figure 1) it is not surprising that their results fit the values reported in this paper. We suggest that for the area around our four drill sites NUS07-3, NUS07-4, NUS07-6, and NUS07-8, the consistently higher values from *Arthern et al.* [2006] are due to the fact that their observational data do not cover the same time period and that in the vicinity of our drill sites no data have been available for validation before. Thus, the interpolation scheme applied by *Arthern et al.* [2006] relies on the comparatively higher accumulation values available for their study.

[30] Even though the time periods are not consistent the differences between our results and the outcomes of studies like those of *Monaghan et al.* [2006] and *Arthern et al.* [2006] stress the importance of ground-based estimates for the validation of large-scale approaches. This is of particular importance in scarcely covered areas like this part of the East Antarctic plateau where mostly older, less reliable data have been available so far [*Magand et al.*, 2007]. While it is not possible to integrate our results in e.g. the study by *Monaghan et al.* [2006] due to the different time periods covered, our study reveals the potential for using ice core data from the remote East Antarctic plateau in addition to carefully selected data from e.g. SPQMLT [*Picciotto et al.*, 1971] to assess accumulation patterns in this area. This will enable further and more detailed studies like the work by *Monaghan et al.* [2006] and *Arthern et al.* [2006] and narrow the error margin on the continent-wide accumulation trend which is important for quantifying the impact of Antarctic mass balance on global sea-level change [*Vaughan*, 2005; *Alley et al.*, 2005].

5. Conclusions

[31] Accumulation from the four firn cores NUS07-3, NUS07-4, NUS07-6 and NUS07-8 is somewhat lower than we would have expected from the SPQMLT data, indicating that parts of the high East Antarctic plateau might have received less precipitation over the last 200 years than previously assumed. Accumulation rates have been approximately stable since the Tambora eruption with values for the period 1815–2007 in the range of 16 to 32 kg m^{-2} a^{-1} with a mean error of 2.3% for that particular time period. The average error for the entire accumulation data set is estimated to be 8%. Comparing the period 1641–1815 with the time span 1815–2007 in the three cores that extend to 1641 (NUS07-3, NUS07-4 and NUS07-6) reveals a decrease of more than 20% in accumulation for the latter period. Some other studies [*Mosley-Thompson et al.*, 1999; *Hofstede et al.*, 2004] have found an increase in accumulation on the high plateau of DML in the (late) 20th century. However, a very recent trend might easily be averaged out in our data. Concerning spatial variability, the first three cores show a decreasing trend whereas core NUS07-8 has a higher accumulation. Clearly, accumulation is anticorrelated with elevation in our data sets. The variation in depth of tracked GPR layers suggests that our firn-core sites are representative of a larger area, yet comparison with SPQMLT results is limited due to different observation periods, large spatial distances and data quality. Our results present insight in spatial and temporal variability of accumulation and contribute new data for this largely uncovered area. More cores and accumulation rates at higher temporal resolution are needed to address questions related to changes in accumulation over the 20th century in this area of the East Antarctic interior.

[32] **Acknowledgments.** This work has been carried out under the umbrella of TASTE-IDEA within the framework of IPY project 152 funded by Norwegian Polar Institute, the Research Council of Norway, and the National Science Foundation of the United States. This work is also a contribution to ITASE. Sanja Forsström and Ryan Banta, Ross Edwards, Dan Pasteris, and Tommy Cox are gratefully acknowledged for help in the lab. Help with Figure 1 was provided by Stein Tronstad and Anders Skoglund. Special thanks goes to the traverse team 2007/2008. Comments by three reviewers significantly improved the manuscript.

References

Alley, R., P. Clark, P. Huybrechts, and I. Joughin (2005), Ice-sheet and sea-level changes, *Science*, *310*(5747), 456–460.

Anschütz, H., O. Eisen, H. Oerter, D. Steinhage, and M. Scheinert (2007), Investigating small-scale variations of the recent accumulation rate in Central Dronning Maud Land, East Antarctica, *Ann. Glaciol.*, *46*, 14–21.

Anschütz, H., D. Steinhage, O. Eisen, H. Oerter, M. Horwath, and U. Ruth (2008), Small-scale spatio-temporal characteristics of accumulation rates in Western Dronning Maud Land, Antarctica, *J. Glaciol.*, *54*(185), 315–323.

Arthern, R., D. Winebrenner, and D. Vaughan (2006), Antarctic snow accumulation mapped using polarization of 4.3-cm wavelength emission, *J. Geophys. Res.*, *111*, D06107, doi:10.1029/2004JD005667.

Budner, D., and J. Cole-Dai (2003), The number and magnitude of large explosive volcanic eruptions between 904 and 1865 A.D.: Quantitative evidence from a new South Pole ice core, in *Volcanism and the Earth's Atmosphere*, *Geophys. Monogr. Ser.*, vol. 139, edited by A. Robock and C. Oppenheimer, pp. 165–176AGU, Washington, D. C.

Cameron, R., E. Picciotto, H. Kane, and J. Gliozzi (1968), Glaciology on the Queen Maud Land traverse, 1964–65, *Inst. Polar Stud. Rep. 23*, 136 pp., Ohio State Univ., Columbus.

Chen, J., C. Wilson, D. Blankenship, and B. Tapley (2006), Antarctic mass rates from GRACE, *Geophys. Res. Lett.*, *33*, L11502, doi:10.1029/2006GL026369.

Cole-Dai, J., E. Mosley-Thompson, S. Wright, and L. Thompson (2000), A 4100-year record of explosive volcanism from an East Antarctica ice core, *J. Geophys. Res.*, *105*(D19), 24,431–24,441.

Davis, C., Y. Li, J. McConnell, M. Frey, and E. Hanna (2005), Snowfall-driven growth in East Antarctic ice sheet mitigates recent sea-level rise, *Science*, *308*(5730), 1898–1901.

Eisen, O., W. Rack, U. Nixdorf, and F. Wilhelms (2005), Characteristics of accumulation rate in the vicinity of the EPICA deep-drilling site in Dronning Maud Land, Antarctica, *Ann. Glaciol.*, *41*, 41–46.

Eisen, O., et al. (2008), Ground-based measurements of spatial and temporal variability of snow accumulation in East Antarctica, *Rev. Geophys.*, *46*, RG2001, doi:10.1029/2006RG000218.

Frezzotti, M., et al. (2004), New estimations of precipitation and surface sublimation in East Antarctica from snow accumulation measurements, *Clim. Dyn.*, *23*, 803–813.

Frezzotti, M., et al. (2005), Spatial and temporal variability of snow accumulation in East Antarctica from traverse data, *J. Glaciol.*, *51*(172), 113–124.

Frezzotti, M., S. Urbini, M. Proposito, C. Scarchilli, and S. Gandolfi (2007), Spatial and temporal variability of surface mass balance near Talos Dome, East Antarctica, *J. Geophys. Res.*, *112*, F02032, doi:10.1029/2006JF000638.

Genthon, C., O. Magnand, G. Krinner, and M. Fily (2009), Do climate models underestimate snow accumulation on the Antarctic Plateau? A re-evaluation of/from in situ observations in East Wilkes Land and Victoria Land, *Ann. Glaciol.*, *50*, 61–65.

Giovinetto, M., and H. Zwally (2000), Spatial distribution of net surface mass accumulation on the Antarctic ice sheet, *Ann. Glaciol.*, *31*, 171–178.

Glen, J., and J. Paren (1975), The electrical properties of snow and ice, *J. Glaciol.*, *15*(73), 15–38.

Hamran, S.-E., and K. Langley (2006), RC-band polarimetric GPR, paper presented at 11th International Conference on Ground Penetrating Radar, Ohio State Univ., Columbus.

Hawley, R., E. Morris, and J. McConnell (2008), Rapid techniques for determining annual accumulation applied at Summit, Greenland, *J. Glaciol.*, *54*(188), 839–845.

Helsen, M., M. van den Broeke, R. van de Wal, W. van de Berg, E. van Meijgaard, C. Davis, Y. Li, and I. Goodwin (2008), Elevation changes in Antarctica mainly determined by accumulation variability, *Science*, *320*(5883), 1626–1629.

Hofstede, C., et al. (2004), Firn accumulation records for the past 1000 years on the basis of dielectric profiling of six firn cores from Dronning Maud Land, Antarctica, *J. Glaciol.*, *50*(169), 279–291.

Hori, A., et al. (1999), A detailed density profile of the Dome Fuji (Antarctica) shallow ice core by X-ray transmission method, *Ann. Glaciol.*, *29*, 211–241.

Intergovernmental Panel on Climate Change (IPCC) (2007), *Climate Change 2007: Synthesis Report. Contribution of Working Groups I, II and III to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, edited by R. K. Pachauri and A. Reisinger, 104 pp., Geneva, Switzerland.

Isaksson, E., and W. Karlen (1994), Spatial and temporal patterns in snow accumulation, western Dronning Maud Land, Antarctica, *J. Glaciol.*, *40*(135), 399–409.

Isaksson, E., W. Karlen, N. Gundestrup, P. Mayewski, S. Whitlow, and M. Twickler (1996), A century of accumulation and temperature changes in Dronning Maud Land, Antarctica, *J. Geophys. Res.*, *101*(D3), 7085–7094.

Isaksson, E., M. van den Broeke, J.-G. Winther, L. Karlöf, J. Pinglot, and N. Gundestrup (1999), Accumulation and proxy-temperature variability in Dronning Maud Land, Antarctica, determined from shallow firn cores, *Ann. Glaciol.*, *29*, 17–22.

Kameda, T., H. Motoyama, S. Fujita, and S. Takahashi (2008), Temporal and spatial variability of surface mass balance at Dome Fuji, East Antarctica, by the stake method from 1995 to 2006, *J. Glaciol.*, *54*(184), 107–116.

Karlöf, L., et al. (2000), A 1500 years record of accumulation at Amundsenisen, Western Dronning Maud Land, Antarctica, derived from electrical and radioactive measurements on an 120 m ice core, *J. Geophys. Res.*, *105*(D10), 12,471–12,483.

Karlöf, L., et al. (2005), Accumulation variability over a small area in East Dronning Maud Land, Antarctica, as determined from shallow firn cores and snow pits: Some implications for ice-core records, *J. Glaciol.*, *51*(174), 343–352.

Koerner, R. (1971), A stratigraphic method of determining the snow accumulation rate at Plateau Station, Antarctica, and application to South Pole Queen Maud Land Traverse 2, 1965–1966, in *Antarctic Snow and Ice Studies II*, *Antarct. Res. Ser.*, vol. 16, edited by A. P. Crary, pp. 225–238, AGU, Washington, D. C.

Langway, C., K. Osada, H. Clausen, C. Hammer, and H. Shoji (1995), A 10-century comparison of prominent bipolar volcanic events in ice cores, *J. Geophys. Res.*, *100*(D8), 16,241–16,247.

Legrand, M., and R. Delmas (1987), A 220-year continuous record of volcanic H_2SO_4 in the Antarctic ice sheet, *Nature*, *327*, 671–676.

Looyenga, H. (1965), Dielectric constants of heterogeneous mixtures, *Physica*, *31*(3), 401–406.

Magand, O., C. Genthon, M. Fily, G. Krinner, G. Picard, M. Frezzotti, and A. Ekaykin (2007), An up-to-date quality-controlled surface mass balance data set for the 90°–180°E Antarctica sector and 1950–2005 period, *J. Geophys. Res.*, *112*, D12106, doi:10.1029/2006JD007691.

McConnell, J., G. Lamorey, S. Lambert, and K. Taylor (2002), Continuous ice-core chemical analyses using Inductively Coupled Plasma Mass Spectrometry, *Environ. Sci. Technol.*, *36*, 7–11.

Monaghan, A., et al. (2006), Insignificant change in Antarctic snowfall since the International Geophysical Year, *Science*, *313*(5788), 827–831.

Moore, J., R. Mulvaney, and J. Paren (1989), Dielectric stratigraphy of ice: A new technique for determining total ionic concentrations in polar ice cores, *Geophys. Res. Lett.*, *16*(10), 1177–1180.

Moore, J., H. Narita, and N. Maeno (1991), A continuous 770-year record of volcanic activity from East Antarctica, *J. Geophys. Res.*, *96*(D9), 17,353–17,359.

Mosley-Thompson, E. (1996), Holocene climate changes recorded in an East Antarctica ice core, *NATO ASI Ser., Ser. 1*, *41*, 263–279.

Mosley-Thompson, E., J. Paskievitch, M. Gow, and L. Thompson (1999), Late 20th century increase in South Pole snow accumulation, *J. Geophys. Res.*, *104*(D4), 3877–3886.

Müller, K., A. Sinisalo, H. Anschütz, S.-E. Hamran, J.-O. Hagen, J. R. McConnell, and D. R. Pasteris (2009), An 860 km surface mass-balance profile on the East Antarctic Pleateau derived by GPR, *Ann. Glaciol.*, in press.

Oerter, H., W. Graf, F. Wilhelms, A. Minikin, and H. Miller (1999), Accumulation studies on Amundsenisen, Dronning Maud Land, Antarctica, by means of dielectric profiling and stable-isotope measurements: First results from the 1995–96 and 1996–97 field seasons, *Ann. Glaciol.*, *29*, 1–9.

Pälli, A., J. Kohler, E. Isaksson, J. Moore, J. F. Pinglot, V. Pohjola, and H. Samuelsson (2002), Spatial and temporal variability of snow accumulation using ground-penetrating radar and ice cores on a Svalbard glacier, *J. Glaciol.*, *48*(162), 417–424.

Picciotto, E., G. Grozaz, and W. de Breuck (1971), Accumulation on the South Pole Queen Maud Land Traverse, 1964–1968, in *Antarctic Snow and Ice Studies II*, *Antarct. Res. Ser.*, vol. 16, pp. 257–315, AGU, Washington, D. C.

Rahmstorf, S., A. Cazenave, J. Church, J. Hansen, R. Keeling, D. Parker, and R. Somerville (2007), Recent climate observations compared to projections, *Science*, *316*(5825), 709–711.

Richardson, C., and P. Holmlund (1999), Spatial variability at shallow snow-layer depths in central Dronning Maud Land, East Antarctica, *Ann. Glaciol.*, *29*, 10–16.

Richardson, C., E. Aarholt, S. Hamran, P. Holmlund, and E. Isaksson (1997), Spatial distribution of snow in western Dronning Maud Land, East Antarctica, mapped by a ground-based snow radar, *J. Geophys. Res.*, *102*(B9), 20,343–20,353.

Rotschky, G., O. Eisen, U. Nixdorf, and H. Oerter (2004), Spatial distribution of surface mass balance on Amundsenisen plateau, Antarctica, derived from ice-penetrating radar studies, *Ann. Glaciol.*, *39*, 265–270.

Stenni, B., M. Proposito, R. Gragnani, O. Flora, J. Jouzel, S. Falourd, and M. Frezzotti (2002), Eight centuries of volcanic signal and climate change at Talos Dome (East Antarctica), *J. Geophys. Res.*, *107*(D9), 4076, doi:10.1029/2000JD000317.

Takahashi, S., Y. Ageta, Y. Fujii, and O. Watanabe (1994), Surface mass balance in east Dronning Maud Land, Antarctica, observed by Japanese Antarctic Research Expeditions, *Ann. Glaciol.*, *20*, 242–248.

Traufetter, F., H. Oerter, H. Fischer, R. Weller, and H. Miller (2004), Spatio-temporal variability in volcanic sulphate deposition over the past 2 kyr in snow pits and firn cores from Amundsenisen, Antarctica, *J. Glaciol.*, *50*(168), 137–146.

Urbini, S., M. Frezzotti, S. Gandolfi, C. Vincent, C. Scarchilli, L. Vittuari, and M. Fily (2008), Historical behaviour of Dome C and Talos Dome (East Antarctica) as investigated by snow accumulation and ice velocity measurements, *Global Planet. Change*, *60*, 576–588.

van de Berg, W., M. van den Broeke, C. Reijmer, and E. van Meijgaard (2006), Reassessment of the Antarctic surface mass balance using calibrated output of a regional atmospheric climate model, *J. Geophys. Res.*, *111*, D11104, doi:10.1029/2005JD006495.

Vaughan, D. (2005), How does the Antarctic ice sheet affect sea level rise?, *Science*, *308*(5730), 1877–1878.

Vaughan, D., J. Bamber, M. Giovinetto, J. Russell, and A. Cooper (1999), Reassessment of net surface mass balance in Antarctica, *J. Clim.*, *45*(150), 933–946.

Velicogna, I., and J. Wahr (2006), Measurements of time-variable gravity show mass loss in Antarctica, *Science*, *311*(5768), 1754–1756.

Wilhelms, F. (1996), *Leitfhigkeits- und Dichtemessung an Eisbohrkernen*, *Ber. zur Polarforsch.*, vol. 191, Alfred Wegener Inst., Bremerhaven, Germany.

Wilhelms, F., J. Kipfstuhl, H. Miller, H. Heinloth, and J. Firestone (1998), Precise dielectric profiling of ice cores: A new device with improved guarding and its theory, *J. Glaciol.*, *44*(146), 171–174.

Winther, J.-G., et al. (1997), EPICA Dronning Maud Land pre-survey 1996/97, in *Report of the Norwegian Antarctic Research Expedition 1996/97*, pp. 96–117, Norsk Polarinst., Tromsø, Norway.

Winther, J.-G., et al. (2002), European Project for Ice Coring in Antarctica (EPICA)—Nordic traverse in 2000/2001, in *Report of the Norwegian Antarctic Research Expedition 2000–2001*, pp. 18–29, Norsk Polarinst., Tromsø, Norway.

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